

A Verified Estimation of the El Niño Index Niño-3.4 since 1877

LUCIA BUNGE AND ALLAN J. CLARKE

Department of Oceanography, The Florida State University, Tallahassee, Florida

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ABSTRACT

Decadal and longer time-scale variabilities of the best known El Niño–Southern Oscillation (ENSO) indexes are poorly correlated before 1950, and so knowledge of interdecadal variability and trend in ENSO indexes is dubious, especially before 1950. To address this problem, the authors constructed and compared physically related monthly ENSO indexes. The base index was El Niño index Niño-3.4, the sea surface temperature (SST) anomaly averaged over the equatorial box bounded by 5°N, 5°S, 170°W, and 120°W; the authors also constructed indexes based on the nighttime marine air temperature over the Niño-3.4 region (NMAT3.4) and an equatorial Southern Oscillation index (ESOI). The Niño-3.4 index used the “uninterpolated” sea surface temperature data from the Second Hadley Centre Sea Surface Temperature dataset (HadSST2), a dataset with smaller uncertainty and better geographical coverage than others. In constructing the index, data at each point for a given month were weighted to take into account the typical considerable spatial variation of the SST anomaly over the Niño-3.4 box as well as the number of observations at that point for that month. Missing monthly data were interpolated and “noise” was reduced by using the result that Niño-3.4 has essentially the same calendar month amplitude structure every year. This 12-point calendar month structure from April to March was obtained by an EOF analysis over the last 58 yr and then was fitted to the entire monthly time series using a least squares approach. Equivalent procedures were followed for NMAT3.4 and ESOI. The new ESOI uses Darwin atmospheric pressure in the west and is based on theory that allows for variations of the atmospheric boundary layer depth across the Pacific.

The new Niño-3.4 index was compared with NMAT3.4, the new ESOI, and with a record of $\delta^{18}\text{O}$ from a coral at Palmyra, an atoll inside the region Niño-3.4 (Cobb et al.). Correlation coefficients between Niño-3.4 and the three monthly indexes mentioned above before 1950 are 0.84, 0.87, 0.73 and 0.93, 0.86, 0.73 for decadal time scales. These relatively high correlation coefficients between physically related but independent monthly time series suggest that this study has improved knowledge of low-frequency variability. All four indexes are consistent with a rise in Niño-3.4 SST and the weakening of the equatorial Pacific winds since about 1970.

1. Introduction

Among the most widely used El Niño–Southern Oscillation (ENSO) indexes are Niño-3.4, the sea surface temperature (SST) anomaly averaged over the east-central equatorial region 5°N–5°S, 170°–120°W, and the Tahiti minus Darwin Southern Oscillation index (SOI), the surface pressure anomaly at Tahiti divided by its standard deviation minus the surface pressure anomaly at Darwin divided by its standard deviation. A monthly Niño-3.4 index, based on the global 5° × 5° “reduce space optimal estimation” by Kaplan et al. (1998), is

available since 1856, while a continuous Tahiti–Darwin SOI record is available since 1876 (at <ftp://ftp.bom.gov.au>). Fig. 1 shows that, while the decadal and longer variability of these indexes agrees reasonably well after 1950 (correlation $r = 0.93$), they are uncorrelated before 1950 ($r = 0.2$). This large drop in correlation is either due to a real change in the relationship between El Niño and the Southern Oscillation or one or other or both of the time series are in error. Note that (see Table 1)

- 1) this drop in correlation does not change significantly if decadal and longer Niño-3.4 variability is derived instead from other available SST reconstruction datasets: the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST; Rayner et al. 2003) or extended reconstructed SST, version 3 (ERSST V3; Smith et al. 2008).

Corresponding author address: Lucia Bunge, Department of Oceanography, The Florida State University, Tallahassee, FL 32306-4320.
E-mail: lbunge@fsu.edu

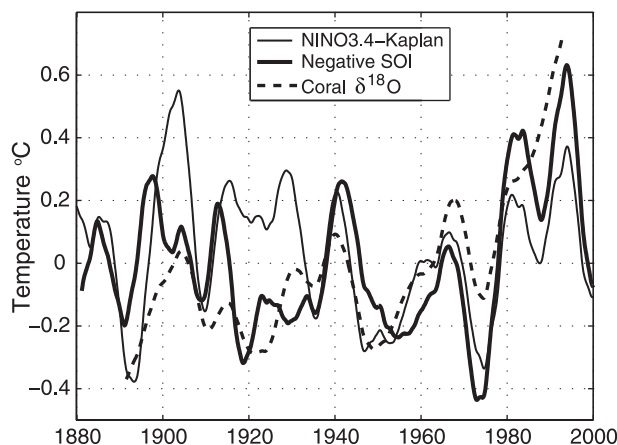


FIG. 1. The Kaplan Niño-3.4 index (thin), the negative SOI (thick), and coral proxy record (Cobb et al. 2003) of Niño-3.4 (dashed). The original monthly time series were filtered with a 75-month running mean followed by a 51-month running mean to isolate decadal and longer variability. The SOI, defined as in the first paragraph of the introduction, was divided by its standard deviation (2.20) and multiplied by 0.22°C , the standard deviation of the Kaplan Niño-3.4 index. In this way the SOI time series is converted into a proxy Niño-3.4 time series in $^{\circ}\text{C}$. Similarly, dividing the $\delta^{18}\text{O}$ record by its standard deviation (0.10‰) and multiplying by the Kaplan Niño-3.4 standard deviation converts the monthly $\delta^{18}\text{O}$ record into a proxy Niño-3.4 record in $^{\circ}\text{C}$. Post-1950, the correlation coefficient between Niño-3.4 and SOI is $r = 0.93$ [$r_{\text{crit}}(95\%) = 0.65$] and pre-1950 $r = 0.20$ [$r_{\text{crit}}(95\%) = 0.53$]. Post-1950, the correlation coefficient between Niño-3.4 and the coral proxy is $r = 0.83$ [$r_{\text{crit}}(95\%) = 0.64$] and pre-1950 $r = 0.55$ [$r_{\text{crit}}(95\%) = 0.57$]. Here and elsewhere r_{crit} is based on Ebisuzaki (1997).

- 2) On the interannual time scale, at the 95% confidence level, the correlations between Niño-3.4 and the SOI are significant both before and after 1950 (see Table 1). However, correlations before 1950 are consistently lower (about 0.7 compared with about 0.9, see Table 1).

Real changes in the relationship between Niño-3.4 and SOI are unlikely because the two time series are connected physically in a fairly simple way. Specifically, since the atmospheric pressure anomalies are large scale, the pressure anomaly at Darwin is similar to the equatorial surface pressure anomaly in the western Pacific, and the surface pressure anomaly at Tahiti is similar to the equatorial surface pressure anomaly in the eastern Pacific. This is true for both interannual (see Figs. 5 and 7 of Deser and Wallace 1990) and decadal (see Fig. 11 of Zhang et al. 1997) time scales. At the equator, the Coriolis parameter is zero and the equatorial surface pressure anomaly difference drives anomalous zonal wind anomalies that tilt the thermocline and cause equatorial SST anomalies in the Niño-3.4 region. Evidence for this mechanism is also available on the

TABLE 1. Correlations (in percent) between the negative SOI and the Niño-3.4 SST index derived from various datasets. The “B” columns in the table refer to correlations before 1950 and the “A” columns to correlations after 1950. “Interannual” time series are the monthly anomalies filtered by 5-month running mean followed by a 3-month running mean. “Decadal and longer” time series refer to time series filtered by a 75-month running mean followed by a 51-month running mean. The numbers in brackets refer to the critical correlation coefficients at the 95% level as calculated using Ebisuzaki (1997). Notice that in all cases correlations among decadal and longer time scales time series are insignificant before 1950 and significant after 1950.

SST dataset	Interannual		Decadal and longer	
	B	A	B	A
Kaplan (Kaplan et al. 1998)	70 (24)	89 (32)	20 (53)	93 (65)
HadISST (Rayner et al. 2003)	69 (24)	90 (29)	16 (54)	90 (65)
ERSST V3 (Smith et al. 2008)	72 (24)	90 (30)	44 (55)	90 (63)

interannual (Fig. 26 of Kessler 1990; Clarke and Lebedev 1996) and decadal time scales (Clarke and Lebedev 1999). These simple physical balances linking the SOI and Niño-3.4 are unlikely to be violated so it is unlikely that the large drop in correlation before 1950 is due to a real physical change.

The above suggests that one or both of the time series are seriously in error before 1950, at least for decadal and lower-frequency variability. Trenberth and Hoar (1997) noted that the Tahiti pressure record is less reliable before 1935 and certainly there are fewer and less quality controlled SST measurements to construct Niño-3.4 before 1950.

One way to test whether it is just one of the time series in error is to correlate both with a third time series that does not suffer from less precise measurements before 1950. Such a record is the long monthly coral proxy record of Niño-3.4 from the Pacific atoll of Palmyra (Cobb et al. 2003). Correlation of this record with Niño-3.4 or SOI at decadal and longer-term variability shows a drop in correlation from post-1950 to pre-1950 in both cases, suggesting that there may be measurement error in both before 1950. Specifically, the coral record correlation with the Niño-3.4 falls from 0.83 after 1950 to 0.55 before 1950 (Fig. 1) while with the SOI it falls from 0.85 after 1950 to 0.43 before 1950.

Although the coral record suggests that both the Niño-3.4 and SOI record may be in error before 1950, this does not imply that the coral record models the long-time-scale variability perfectly. In particular, since 1975, when the Niño-3.4 data are of good quality, the coral record suggests a sharp increase in temperature not seen in Niño-3.4 (Fig. 1).

The above evidence suggests that we do not know the decadal variations and long-term trends in ENSO

indexes very well. Yet, if we are to assess the influence of global warming on equatorial Pacific and related tele-connected climate change, and if we are to assess whether decadal and longer changes in background state can affect the frequency and size of El Niño events, it is vital that we have long and accurate ENSO time series.

Motivated by the above, in this paper we will describe the construction and verification of monthly values of Niño-3.4 since 1877. We chose Niño-3.4 because of the longevity of the raw SST data and their close relationship to key ENSO dynamical variables. Niño-3.4 is, for example, highly correlated (see Fig. 2) with the equatorial sea level and thermocline tilt forced by the interannual winds (Kessler 1990; Li and Clarke 1994), and the time integral of Niño-3.4 is well correlated with another key dynamical ENSO variable, the equatorial warm water volume anomaly (Meinen and McPhaden 2000; Clarke et al. 2007).

We used the improved dataset of Rayner et al. (2006) to construct a new Niño-3.4 index. We also constructed a Niño-3.4 proxy index based on nighttime marine air temperature (NMAT3.4) data over the Niño-3.4 region and a new equatorial SOI (ESOI), which takes into account the depth of the atmospheric boundary layer at each end of the equatorial Pacific. These two extra indexes, both closely related to the variability in the Niño-3.4 region, allow us to verify our new Niño-3.4 index. Results show a better agreement before and after 1950, not only for the time series we constructed but also with the coral Niño-3.4 proxy record. Finally, the trends of the last 30 yr in the indexes suggest that global warming is causing the equatorial Pacific trade winds to weaken and the SST over Niño-3.4 to rise.

The paper is organized as follows. Section 2 describes the construction of the new Niño-3.4 index, section 3 the construction of the proxy Niño-3.4 index based on NMAT, and section 4 the construction of the ESOI. Section 5 verifies the new Niño-3.4 index with the other constructed proxies and with the coral proxy record of Niño-3.4 from the Pacific atoll of Palmyra (Cobb et al. 2003). The verification is done for both interannual and interdecadal time scales. Section 6 discusses low-frequency variability in all our indexes and its relationship to global warming, and section 7 summarizes and discusses our main results.

2. Construction of a monthly Niño-3.4 index

To construct the new Niño-3.4 index, we used an improved SST dataset, HadSST2 (Rayner et al. 2006). The HadSST2 dataset consists of in situ measurements of SST from ships and buoys. The measurements were

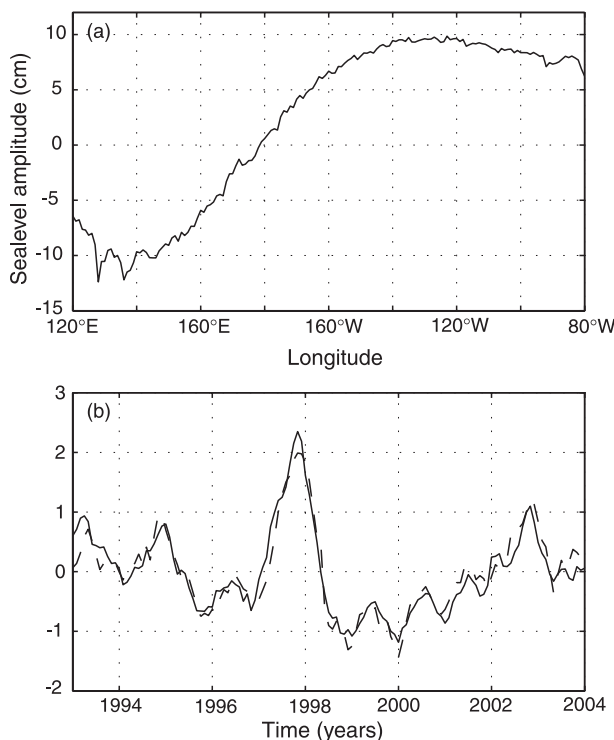


FIG. 2. (a) The first EOF (69% of the variance) of the sea surface height anomaly (in cm) for the equatorial (5°S–5°N) Pacific, from satellite data (<http://las.aviso.oceanobs.com>). (b) The corresponding first principal component (solid line). The principal component time series has been normalized to have variance = 0.5 so that (a) represents the amplitude of the sea level anomaly. The dashed curve in (b) is the Niño-3.4 time series normalized so that it has the same variance as the first principal component. The correlation coefficient between the two time series is $r = 0.95$ (from Clarke et al. 2007).

converted to anomalies by subtracting climatological values and then averaging on a $1^\circ \times 1^\circ$ monthly grid. After gridding the anomalies, the data were corrected for spurious trends caused by changes in SST measuring practices before 1942. Contrary to other SST datasets (HadISST, Rayner et al. 2003; Kaplan, Kaplan et al. 1998; ERSST V3, Smith et al. 2008), which are often used for long-term SST analysis (e.g., Liu et al. 2005; Vecchi et al. 2008), SST data in the HadSST2 dataset have not been interpolated over large regions. Rather, the $1^\circ \times 1^\circ$ monthly estimate in HadSST2 is given by the average of the available observations for that specific grid point, for that month, and, if there are no data, then no value is recorded. As pointed out by Rayner et al. (2003), interpolation over large regions can introduce error, and Niño-3.4 is, unfortunately, a data-sparse region (Fig. 3). In a series of steps below, we explain our attempt to obtain a new Niño-3.4 index that is as reliable as possible.

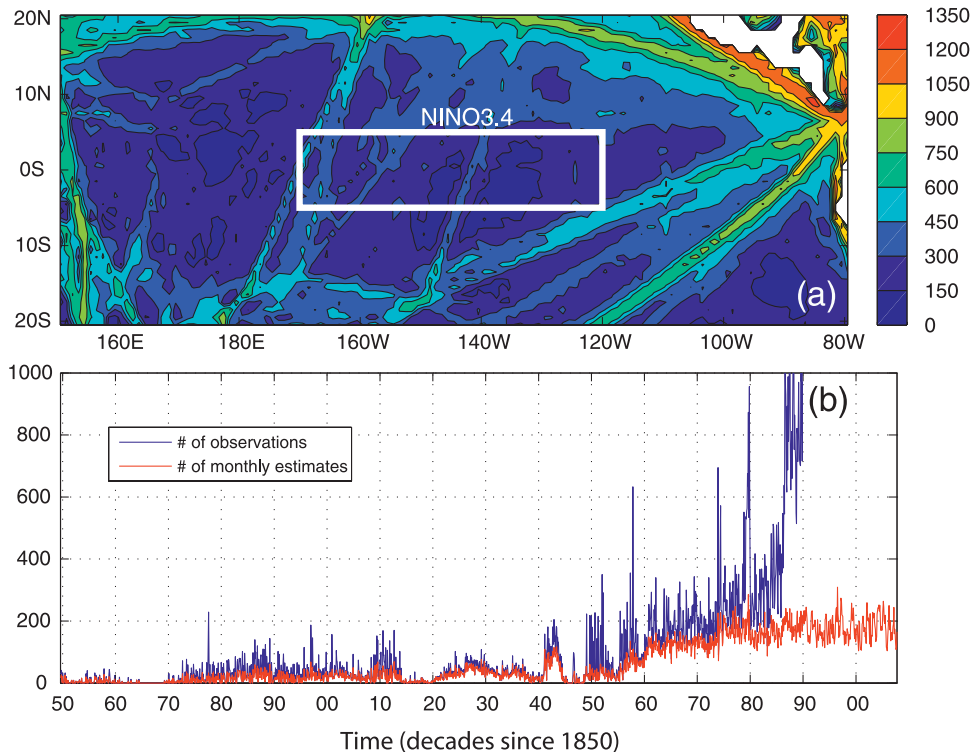


FIG. 3. (a) Number of monthly SST estimates over the tropical Pacific since 1850. The highest densities of monthly SST estimates in region Niño-3.4 (white box) are found in the western side of the box. (b) The number of monthly observations in the Niño-3.4 box (blue) and the number of grid points with monthly estimates in the Niño-3.4 box (red) as a function of time. Critical data gaps occur before 1873, especially in the 1860s where there are many years with no data; around 1917; and in the early and late 1940s. After 1985, with the initiation of the TAO array (McPhaden et al. 1998), the number of observations increased considerably; after 1990, there are more than 1000 observations for each monthly Niño-3.4 index estimate.

a. Step 1: Take into account the uneven distribution of data in space by considering the spatial variations in SST anomaly amplitude over the Niño-3.4 region

An EOF of SST over the Niño-3.4 region (Fig. 4a) shows that the SST anomaly amplitude varies over the Niño-3.4 box. This implies that, when data are scarce, an error in the calculation of Niño-3.4 can be introduced if the observations come mostly from a region of the box where the amplitude differs from the average amplitude over the Niño-3.4 region. From the spatial distribution of data in the region (Fig. 3a), we can see that this is indeed the case: before the Tropical Atmosphere Ocean (TAO) project array starting in 1985 (McPhaden et al. 1998), most of the observations are found at the western part of the box, an area where the amplitudes tend to be smaller than in the rest of the box (Fig. 4). One caveat with this approach is that the horizontal weighting structure of the Niño-3.4 box in Fig. 3a may change decadal. To confirm the structure for earlier years,

we computed the ratio of standard deviations between the individual gridpoint SST anomaly time series in Niño-3.4 and the average time series of Niño-3.4. The horizontal weighting structure obtained in this way is similar to the one obtained by the EOF based on more recent data (Fig. 4b): the largest anomalies are situated on the equator and increase in strength toward the east. There are, however, small differences between the amplitudes of the weighting structures that can be attributable to noise. For that reason, we opted to use the horizontal weighting structure resulting from the EOF.

b. Step 2: Take into account the number of monthly observations at different $1^\circ \times 1^\circ$ Niño-3.4 grid points

There are 500 $1^\circ \times 1^\circ$ grid points in the Niño-3.4 region. The HadSST2 dataset estimate for some grid points for a given month may be based on one or two SST observations, while at another grid point the monthly average might be based on hundreds of observations

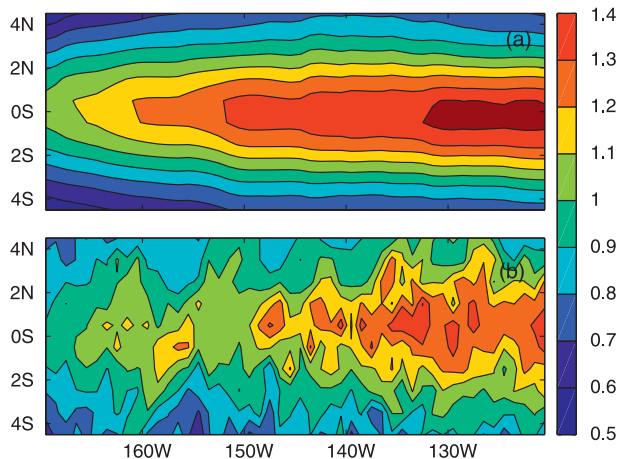


FIG. 4. (a) Amplitude of the first EOF of the SST anomalies in the Niño-3.4 region since 1982. This EOF describes 89% of the variance and its corresponding principal component is correlated with Niño-3.4 with $r = 0.9994$. Because an EOF analysis requires a complete dataset (no gaps), the data used for this analysis are the Reynolds SST data (Reynolds et al. 2002). (b) Ratio of standard deviations between individual time series and the box average time series (HadSST2 data) over the entire period of time (1850–2006). The resulting values were normalized so the average amplitude in the box is equal to 1.

(Fig. 3b). Thus, gridpoint values can vary greatly in their reliability. To overcome this, each gridpoint observation in the Niño-3.4 box for a given month was weighted by the number of observations at the grid point for that month divided by the total number of observations in the Niño-3.4 box for that particular month.

c. Step 3: Take into account too few data in the Niño-3.4 box for a given month

The number of observations and estimates is much smaller before 1985, and there are serious gaps in the 1860s, around 1917, and in the early and late 1940s (Fig. 3b). It is unlikely that original data will become available to fill these gaps, so any past or new estimate of Niño-3.4 is hampered by the same problem. We decided to use all the available data for the construction of the Niño-3.4 index, keeping in mind their limitations when there are few observations. Checking the validity of the index in periods of little data was left to a posterior step in this analysis, namely, during the verification of the Niño-3.4 index with other physically related indexes (section 5).

d. Step 4: Use the 12-month April–March amplitude structure of Niño-3.4 to interpolate missing or doubtful data

One characteristic of ENSO phenomena is the seasonal phase locking of amplitude. For example, Niño-3.4

typically begins to grow in amplitude in boreal spring, reaching an amplitude peak in December and then decreasing in amplitude in March. This can be seen quantitatively in Fig. 5 (Clarke 2008). An EOF analysis of the April–March Niño-3.4 time series shows that the first EOF describes 83% of the variance so that

$$\text{Niño-3.4}(m, t) \approx S(m)Y(t), \quad (2.1)$$

where m is one of the 12 calendar months and t is a given El Niño year. Figure 5c shows that (2.1) is an excellent approximation to Niño-3.4 so that almost always we may think of Niño-3.4 as having the same calendar structure $S(m)$ (Fig. 5a) every year (April–March), the only difference from one year to the next being the value of the annual time series $Y(t)$ (Fig. 5b). This property of ENSO and the Niño-3.4 index offers a way to interpolate missing or doubtful data by fitting the calendar structure $S(m)$ to the monthly time series. Furthermore, fitting the calendar structure $S(m)$ to the series serves as a filter by reducing noise in the series. The reconstruction of the Niño-3.4 index in this way depends on $S(m)$ being constant on decadal and longer time scales. We estimated two $S(m)$ structures for the periods 1896–1906 and 1921–38, where there are no gaps in the data, and compared them with the $S(m)$ structure from the period 1950–2008 (Fig. 6). The maximum discrepancies between the different $S(m)$ structures do not exceed 20% and the correlation coefficients between the time series fitted with the different $S(m)$ structures are above 0.97 for both monthly and decadal filtered time series. Thus, the choice of $S(m)$ from these three periods does not affect the results. We used the amplitude structure $S(m)$ based on the period between 1950 and 2008 because the data quality is higher and the record is the longest (Fig. 5a).

The decadal filtered time series resulting from each step described above are shown in Fig. 7. Weighting by the horizontal structure has little effect (Fig. 7, thin black and thick gray lines). The largest changes, of about 0.1°C , are seen when the $1^{\circ} \times 1^{\circ}$ gridpoint data are weighted by the number of observations, especially after 1985, when the TAO array (see e.g., McPhaden et al. 1998) began taking data. The TAO array provides hourly values of SST, producing a disproportionate number of observations where TAO moorings are located. Thus, after 1985, the signal is predominantly given by the instruments in the TAO array. The nearly systematic difference of 0.1°C between weighted and unweighted time series after 1985 suggests differences in the measurements taken by TAO moorings compared to ships.

The fitting of the calendar-year structure (dashed gray line) does not significantly change interdecadal variability

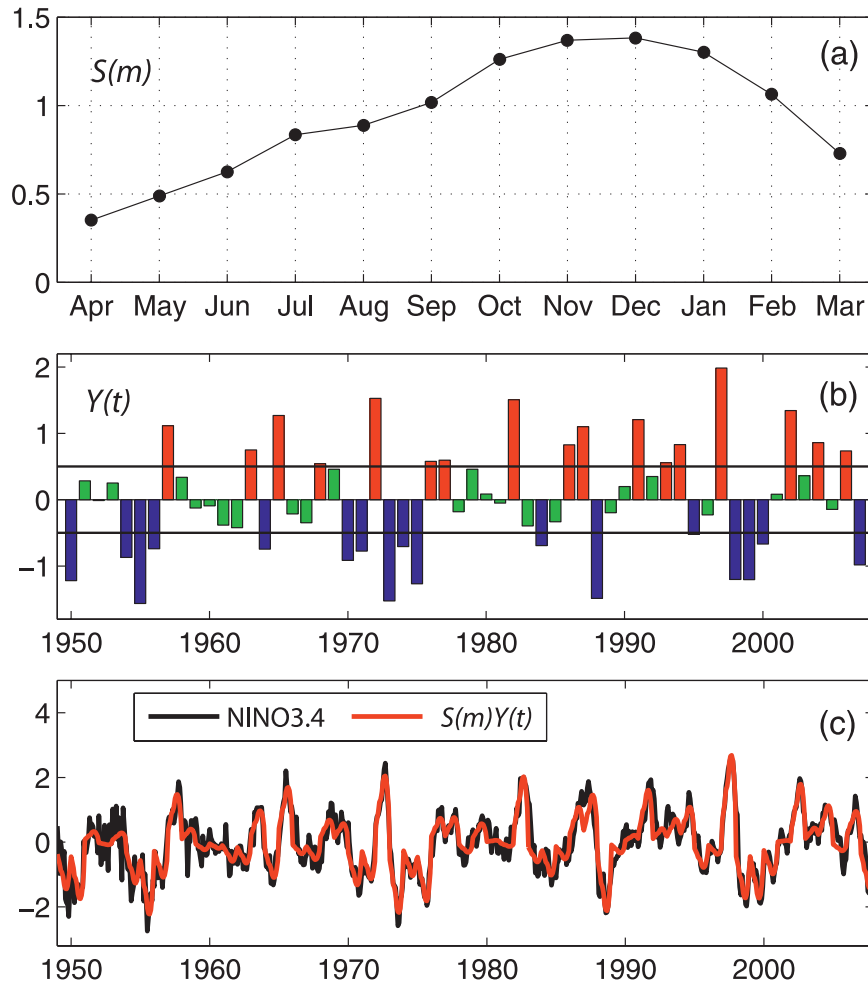


FIG. 5. Representation of the Niño-3.4 index (in $^{\circ}\text{C}$) as the product of $S(m)Y(t)$ where $S(m)$ ($m = 1$ for April, $m = 2$ for May, ..., $m = 12$ for March) is the calendar-year structure function and $Y(t)$ ($t = 1950, 1951, \dots, 2008$) is the annual time series amplitude function. The product $S(m)Y(t)$ was determined as the first mode of an EOF analysis and represents 83% of the Niño-3.4 index variance. (a) The structure function $S(m)$. The structure function has been normalized so that $\sum_{m=1}^{12} [S(m)]^2 = 12$. (b) The annual amplitude function $Y(t)$. When $|S(m)Y(m)|$ exceeds 0.5°C for at least 3 months of the year, the year is an El Niño year ($Y > 0$, red) or a La Niña year ($Y < 0$, blue). When $|S(m)Y(m)| \leq 0.5^{\circ}\text{C}$, the year is neutral (green). (c) Niño-3.4 (black line) and $S(m)Y(t)$ (red line). The correlation coefficient between these two time series is $r = 0.91$ (updated version of Fig. 8.6 from Clarke 2008).

when compared to series from the previous step (dotted black line). However, by filling gaps in months that have no data, calendar-year fitting significantly increases the length of the continuous monthly time series.

3. Construction of a Niño-3.4 proxy index based on NMAT data

Because of the turbulent mixing of air near the surface, one might expect the NMAT overlying the Niño-3.4 region to be closely related to the underlying SST and therefore to Niño-3.4. Figure 3 of Folland et al. (2003)

shows that the correlation between SST and NMAT is greater than 0.9 over the Niño-3.4 region, suggesting that NMAT data can be used to check the Niño-3.4 SST index.

To construct the Niño-3.4 proxy, NMAT3.4, from NMAT data, we use the Met Office historical marine air temperature (MOHMAT43) dataset of Rayner et al. (2003). This dataset is based on in situ measurements from ships and buoys. The measurements were converted to anomalies by subtracting climatological values and then averaged onto a $5^{\circ} \times 5^{\circ}$ monthly grid. After gridding the anomalies, bias corrections were applied to

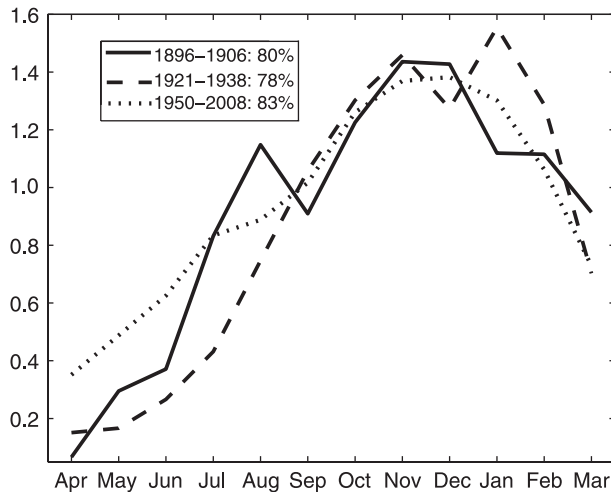


FIG. 6. Calendar-year structure function based on EOFs for the periods 1896–1906 (solid), 1921–38 (dashed), and 1950–2008 (dotted). The structure functions have been normalized so that in each case $\sum_{m=1}^{12} [S(m)]^2 = 12$. The variance explained by these EOFs is 80%, 78%, and 83%, respectively. The periods were chosen based on the intervals of the Niño-3.4 monthly time series that had no gaps.

remove spurious trends caused by changes in ship deck heights and various unusual operational practices. Compared to the HadSST2 dataset, this dataset has much lower spatial resolution ($5^\circ \times 5^\circ$ instead of $1^\circ \times 1^\circ$) and the number of observations used to calculate the monthly estimate at each grid point is not available. These two conditions reduce the resolution of the horizontal weighting structure and do not allow weighting by the number of observations at each grid point. Furthermore, the times when NMAT3.4 data are scarce or nonexistent are the same as when SST data are scarce or nonexistent. However, the NMAT data are independent measurements and so can confirm the variability where data exist and also provide information about trends.

NMAT data were weighted by a horizontal amplitude structure resulting from the ratio of standard deviations between the average time series of the Niño-3.4 box and the individual time series in the Niño-3.4 box. As for the SST case, a calendar structure function $S(m)$ was obtained from an EOF analysis of the time series since 1950 and fitted to the data. The obvious outliers, present in periods of scarce data, were removed. The resulting time series is discussed in section 5.

4. Construction of a Niño-3.4 proxy index based on surface atmospheric pressure difference between the eastern and western equatorial Pacific (ESOI)

In this section we extend the ESOI work of Clarke and Lebedev (1996). We generalize the ESOI so that it

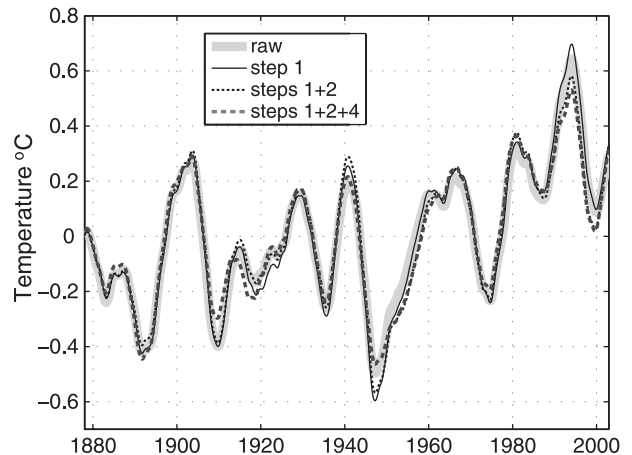


FIG. 7. Decadally filtered time series of the original monthly Niño-3.4 (thick light gray line) after weighting the series by the horizontal structure of Niño-3.4 (thin black line), adding the weighting by the number of observations (dotted black line), and adding the fitting of the calendar-month structure function $S(m)$ (dashed gray line). The monthly time series of the dashed gray line is complete and was decadal filtered in the same way as Fig. 1. The other 3 time series had data gaps and were filtered as follows: if there were, say, n months missing for a given 75-month time interval, the gaps were ignored and the sum of $75 - n$ data points was divided by $75 - n$. The monthly data were then filtered by a 51-month running mean as in Fig. 1.

allows for a surface equatorial atmospheric boundary layer that can vary in depth across the Pacific and we extend the ESOI time series back to 1877.

a. Theory

At the equator, the Coriolis parameter is zero and a difference in anomalous surface pressure will drive anomalous zonal wind anomalies. These anomalies drive anomalous ocean flow, which generates SST anomalies in the Niño-3.4 region. The linear relation between zonal wind stress anomalies and equatorial surface atmospheric pressure difference across the equatorial Pacific was established by Clarke and Lebedev (1996). In the following, we will extend their analysis to allow for an atmospheric boundary layer that can vary in depth across the Pacific.

Consider the zonal momentum equation in the atmosphere:

$$\frac{\partial u}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{u} - f v = -\frac{p_x}{\rho} + \frac{X_z}{\rho}. \quad (4.1)$$

In this equation x , y , and z refer to eastward, northward, and upward coordinates; $\mathbf{u}$ is the three-dimensional velocity vector (u , v , w); p is the pressure; ρ is the air density; ∇ is the three-dimensional gradient operator; f is the Coriolis parameter; and X is the eastward turbulent

stress due to friction with the ocean surface. On the interannual and interdecadal time scales of interest, $\partial u/\partial t$ is negligible compared to p_x/ρ . The Coriolis term can also be dropped at the equator where f is zero and, to a first approximation (Clarke and Lebedev 1996), we can neglect nonlinear advection terms. Equation (4.1) then reduces to

$$p_x = X_z. \quad (4.2)$$

Integration over the boundary layer of height H gives

$$\int_0^H p_x dz = -\tau^x, \quad (4.3)$$

where τ^x is the eastward surface wind stress. If p_x were independent of z in the boundary layer, then the left-hand side of (4.3) would be Hp_x . However, particularly in the eastern equatorial Pacific, p_x may decrease from the surface to the top of the boundary layer. In that case

$$\int_0^H p_x dz = hp_x^{\text{surface}}, \quad \text{with } h < H. \quad (4.4)$$

Combining (4.3) and (4.4) we have

$$hp_x = -\tau^x, \quad (4.5)$$

where, for notational convenience, p refers to the surface pressure anomaly. In (4.5), h is a function of x and is expected to be larger in the west than in the east because the surface stress layer is probably deeper in the west, due to the higher SST and less stable overlying air. Therefore, h decreases eastward.

Integrating (4.5) zonally across the Pacific from the western boundary $x = 0$ to the eastern boundary $x = L$ gives

$$\int_0^L hp_x dx = -\int_0^L \tau^x dx, \quad (4.6)$$

or, upon integration by parts,

$$\int_0^L hp_x dx = [hp]_0^L - \int_0^L h_x p dx. \quad (4.7)$$

The term $\int_0^L h_x p dx$ is likely to be much smaller than $h(0)p(0)$ because h_x is of one sign and p changes sign across the Pacific, being positive and negative by about the same amount. Therefore,

$$\int_0^L hp_x dx \approx h(L)p(L) - h(0)p(0). \quad (4.8)$$

From (4.6) and (4.8) we can write

$$p(L) - \frac{h(0)}{h(L)}p(0) = \frac{-1}{h(L)} \int_0^L \tau^x dx. \quad (4.9)$$

Based on the above we define a new ESOI by

$$\text{ESOI} = p(L) - \frac{h(0)}{h(L)}p(0). \quad (4.10)$$

The subsections below will discuss the construction of $p(L)$ and $p(0)$ and the estimation of $h(0)/h(L)$. Based on theory, we expect $h(0)/h(L)$ to be greater than one; that is, in the ESOI, western equatorial Pacific surface air pressure is weighted more than eastern equatorial surface air pressure.

b. Construction of SLP time series in the east and west Pacific [$p(L)$ and $p(0)$]

In the east, $p(L)$ was constructed by averaging sea level pressure (SLP) anomalies from the International Comprehensive Ocean–Atmosphere Data Set (ICOADS; additional information is available online at <http://icoads.noaa.gov/>). The ICOADS SLP dataset consists of monthly summary statistics of observations taken primarily from ships but also from moored and drifting buoys. The SLP data were converted to anomalies by subtracting a climatology made from the records since 1950 to the most recently available data. Because we wanted to have a time series as long and complete as possible for the SLP in the eastern Pacific (SLPE), we expanded the Clarke and Lebedev (1996) eastern Pacific data box to 3°N–3°S, 120°–83°W. Note that SLP data are also available at 80°W, and that it is along this longitude that the number of data is greatest. However, we found that time series at this longitude had very high amplitudes from the periods of 1916–18 and 1942–45 when data were scarce. These data were also inconsistent with the Darwin SLP time series and for these reasons we ignored them and ended our box at 83°W.

As for the new Niño-3.4 index in section 2, the SLPE raw data were weighted by a horizontal structure estimated using the first mode of an EOF analysis. To calculate this EOF, we first had to obtain complete monthly time series at each box grid point. We filled in missing data at each grid point using the 12-month calendar structure function described in step 4 of section 2. We estimated this function from a complete monthly SLP anomaly time series for the box from a simple box average for each month for the period 1950–2005.

Once the raw gridpoint data were approximately weighted horizontally according to the EOF, each

monthly grid point was weighted according to the number of observations to form a monthly SLPE time series. This time series was used to obtain a second calendar structure function ($S(m)$), which in turn was fitted to the time series, giving a final SLPE time series of the form $S(m)Y(t)$ [see (2.1)].

For the sea level pressure in the western Pacific (SLPW), we initially constructed a time series also based on the ICOADS dataset following equivalent steps to those used to construct SLPE. When compared to the SLP record at Darwin, a long and well-reputed time series reflecting the SLP variability in the western equatorial Pacific, correlation coefficients were relatively high after 1950 [$r = 0.84$, $r_{\text{crit}}(95\%) = 0.31$; here and elsewhere r_{crit} is based on Ebisuzaki (1997)] but dropped before 1950 [$r = 0.53$, $r_{\text{crit}}(95\%) = 0.18$]. Furthermore, there was a better negative correlation between SLPE and Darwin ($r = -0.66$, $r_{\text{crit}} = 0.12$) than between SLPE and SLPW ($r = -0.52$, $r_{\text{crit}} = 0.16$). These two factors made us opt for the Darwin record as the best estimate of western SLP variability.

The SLP Darwin record was downloaded from <http://www.cgd.ucar.edu> and consists of monthly estimates since 1866. The SLP data were converted to anomalies by subtracting a climatology made from the records from 1950 to 2006. To obtain the right amplitude of the signal at the equator for use in the construction of the ESOI, we calculated the ratio of standard deviations between SLPW and Darwin for the time when the series were well correlated. This value is 0.63, implying that Darwin SLP has larger amplitudes than equatorial SLP. Why should this be the case? During El Niño, surface winds between Darwin and the equator are easterly (see Fig. 7 of Deser and Wallace 1990) and the pressure in the western Pacific is anomalously high. Therefore, by geostrophy, the pressure at Darwin must be higher (and therefore larger in amplitude) than the pressure at the equator. During La Niña, the surface winds are westerly, Darwin pressure is anomalously low, and, by geostrophy, lower (and therefore larger in amplitude) than the surface pressure at the equator.

c. Construction of the pressure difference

To calculate ESOI from (4.10), we need to find $h(0)/h(L)$, which we expect to be similar to $H(0)/H(L)$, the ratio between the depth of the atmospheric boundary layer in the west and the depth of the atmospheric boundary layer in the east. The ratio $h(0)/h(L)$ can be calculated from (4.9) using the SLP time series already described and zonal wind stress data. Zonal wind stress (τ^x) since 1978 was obtained from The Florida State University (FSU) winds (<http://www.coaps.fsu.edu>). We

chose a box bounded by 3°N, 3°S, 152°E, and 122°W, the region where variations in the wind can be related to the difference in the pressure difference between our SLP time series. The zonal wind stress anomalies over this box were first averaged meridionally and then integrated over longitude. To reduce noise, a calendar structure function $S(m)$, analogous to the one described in step 4 in section 2, was calculated from the series and then fitted to the series to obtain a zonally integrated zonal wind stress time series of the form $S(m)Y(t)$ [see (2.1)]. Past work (e.g., Ramage 1984, 1987; Peterson and Hasse 1987; Cardone et al. 1990; Clarke and Lebedev 1996) has shown that there are false trends in the wind data because of changes in wind estimation techniques. Therefore, after fitting the time series, we removed the linear trends of both wind stress and pressure before estimating $h(0)/h(L)$. This trend removal was only done for this calculation alone. To calculate $h(0)/h(L)$, we re-wrote (4.9) in the form

$$p(L, t) - \alpha p(0, t) = -\frac{a}{h_*(L)} \int_0^L \tau^x dx \quad (4.11)$$

and calculated a series of regressions of the time series $p(L, t) - \alpha p(0, t)$ for $\alpha = 0.50, 0.51, \dots, 3.99, 4.00$, against the time series $-[a/h_*(L)] \int_0^L \tau^x dx$ with $h(L) = 400$ m and “ a ” the order one dimensionless regression coefficient to be determined. The value of α giving the best correlation coefficient ($r = 0.91$) and establishing the value of $h(0)/h(L)$ to be used in the ESOI was 1.67 (Fig. 8). The regression coefficient “ a ” defined our estimate of $h(L)$ in meters as $h(L) = h(L)/a = 400/a$ and also our estimate of $h(0)$ as $h(0) = \alpha 400/a$. We estimated the coefficient “ a ” in two different ways and obtained similar results:

- 1) By a least squares fit: $a = 1.93$; $h(L) \approx 207$ m; $h(0) \approx 346$ m.
- 2) By the ratio of the standard deviation of the two time series: $a = 2.13$; $h(L) \approx 188$ m; $h(0) \approx 314$ m.

The values for $h(0)$ and $h(L)$ are smaller than observed boundary layer thicknesses $H(0)$ and $H(L)$. For example, Zeng et al. (2004) observed values of $H(L)$ of around 500 m. We expect $h(L) < H(L)$ [see (4.4)], but the difference between $h(L)$ and $H(L)$, if accurate, would imply a strong decrease in p_x in the boundary layer.

In any case, it is the parameter α that is important for the derivation of the ESOI. Consistent with the idea that ESOI, the zonally integrated wind stress, and Niño-3.4 should closely resemble each other, we checked the value of α by replacing, on the right-hand side of (4.11), the zonally integrated wind stress with the Niño-3.4 index derived in section 2. Considering the same interval of

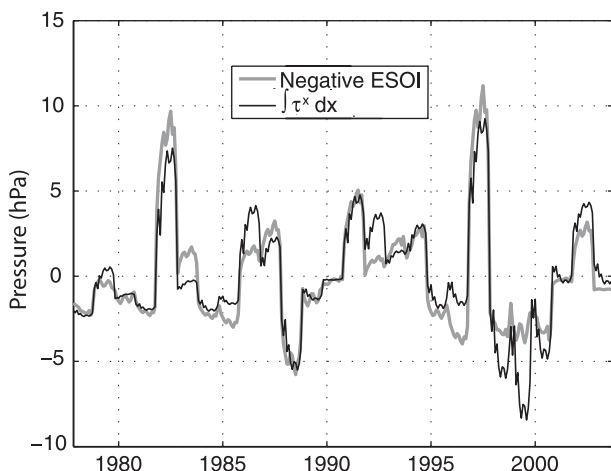


FIG. 8. Negative ESOI calculated as described in section 4, using $h(0)/h(L) = 1.67$ (gray solid line), compared to the zonally integrated zonal wind stress at the equator (black thin line). The two series are correlated at $r = 0.91$ with $r_{\text{crit}}(95\%) = 0.46$.

time for when the wind data are available, we calculate $\alpha = 1.52$. This is within 10% of the value 1.67 found for the wind stress regression. The Niño-3.4/pressure estimation of $\alpha = 1.52$ was for Niño-3.4 and pressure time series that were not detrended since, unlike the wind, we do not expect false trends for these time series. If these time series are detrended, then $\alpha = 1.50$, which is negligibly different from $\alpha = 1.52$. In any case, since we want to keep the ESOI analysis independent of the Niño-3.4 analysis, we chose the wind stress regression value $\alpha = 1.67$ in our definition of the ESOI.

5. Verification

In sections 2 through 4 we constructed three new indexes that are closely related physically. In this section we compare these three indexes with each other and also with the Niño-3.4 coral proxy record from Cobb et al. (2003). If these indexes, which are based on independent datasets, agree well with each other, it is likely that we have captured the climate signal.

We compare the various indexes using correlation coefficients for the entire time series and for the pre- and post-1950 portions of those time series. Correlations were done separately for the monthly time series, which are dominated by the interannual time scale, and also for decadal and longer time scales (Table 2). All correlations are high and significant before and after 1950 for both interannual and decadal and longer time scales. We note that the correlation coefficients are typically as high for decadal and longer time scales as for interannual time scales even though the amplitude

of the decadal and longer signal is only about 10%–20% of the interannual signal (see Fig. 9).

The good correlation between the new Niño-3.4 and NMAT3.4 at interannual as well as at decadal time scales, for the times when NMAT3.4 is available, supports the validity of our Niño-3.4 index. But it is the good correlation between the continuous monthly new Niño-3.4 and new ESOI (Fig. 9) that we find to be the most compelling, because for these two datasets we can be sure that the time of collection of the data as well as the location is completely independent. For the interannual time series, the largest differences between the Niño-3.4 and the ESOI are before 1900, around the beginning of World War I in 1914, around the end of the World War II in 1945, and around the two major El Niño events of 1982 and 1997. In the first three cases the differences are probably due to the scarcity of data (Fig. 3). The differences in the last two cases, during the major El Niño events of 1982 and 1997, are expected because of the size of the events and the different maxima and minima in the $S(m)$ structure (i.e., Niño-3.4 peaks in December while ESOI peaks in February); the percentage error in these large-amplitude cases is not unusually large. For decadal time scales, the largest differences between the new Niño-3.4 and new ESOI occur around 1895, around World War I in 1914, and in the 1960s. In the first two cases, as with interannual variability, the differences can be attributed to the scarcity of data in both datasets. We do not know the reason for the differences in the 1960s, particularly because it is a period where differences in monthly time series are small.

Regardless of the differences between indexes, all indexes constructed here are much better correlated on decadal and longer time scales before 1950 than the indexes of Kaplan Niño-3.4 or the SOI (Fig. 1 and Tables 1 and 2). These results suggest that the new indexes provide a more accurate knowledge of decadal and interdecadal variability in the last 130 yr.

Specifically, how does the new Niño-3.4 index differ from previous Niño-3.4 indexes on decadal and longer time scales? Fig. 10 shows decadal filtered time series of the new Niño-3.4 index together with Kaplan-, HadISST-, and ERSST V3-derived Niño-3.4 indexes. The 4 indexes show very similar quasi-decadal variability but there are sizeable differences between the average values of the indexes before 1955. These differences translate into different general trends (Table 3). For the years in question, all four indexes essentially use the same dataset, namely, the Comprehensive Ocean–Atmosphere Data Set (COADS). We therefore conclude that the differences in the indexes are due to the interpolation schemes used in the SST reconstructions.

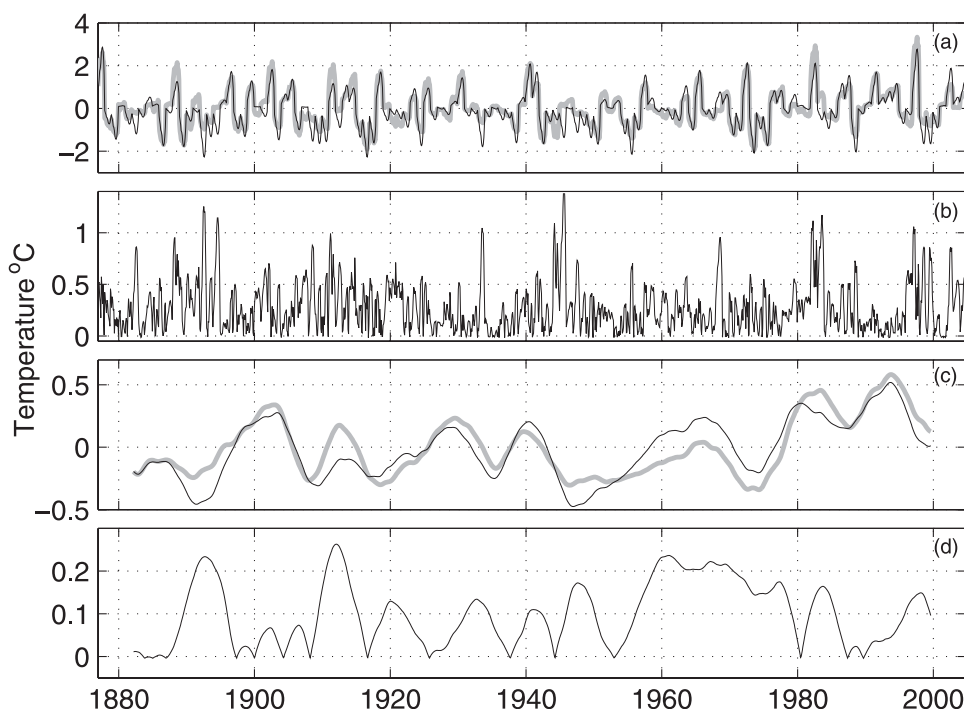


FIG. 9. (a) Monthly indexes of Niño-3.4 (black, in $^{\circ}\text{C}$) and negative ESOI (gray, in hPa). The ESOI has been multiplied by 0.66, the ratio of the standard deviation of Niño-3.4 (0.84°C) to the standard deviation of the ESOI (1.28 hPa). In this way the ESOI is converted into a proxy Niño-3.4 time series. (b) Absolute value of the difference between monthly Niño-3.4 and negative ESOI. (c) Same as (a) but for decadal time scales. In this case the ratio of standard deviations is 0.71, and the standard deviations of Niño-3.4 and the ESOI are 0.23°C and 0.32 hPa. (d) Same as (b) but for decadal time scales.

6. Niño-3.4 index and ESOI in the context of global warming

Over the last decade, there has been considerable debate (Vecchi et al. 2008) over the response of the tropical Pacific to increasing greenhouse gas concentration in the atmosphere. One theory, the “ocean thermostat” mechanism (Clement et al. 1996; Cane et al. 1997),

based on the Zebiak–Cane ocean–atmosphere couple model (Zebiak and Cane 1987), predicts a La Niña–like response. Specifically, the easterly winds and the thermocline tilt across the Pacific intensify, and the SST in the eastern equatorial Pacific decreases. On the other hand, Vecchi and Soden (2007) analyzed the response of the tropical atmospheric and oceanic circulation to greenhouse gases of 22 model experiments of the

TABLE 2. Correlation coefficients (in percent) between the three indexes constructed in sections 2–4 (Niño-3.4, NMA3.4, and negative ESOI) and the coral Niño-3.4 proxy record from Palmyra (Cobb et al. 2003). All correlation values are significant at a 95% confidence based on Ebisuzaki (1997). The nomenclature in the table means the following: B = before 1950; A = after 1950; T = entire time series. In each case the length of the time series in the correlations was determined by the shortest time series. For interannual time scales, Niño-3.4, NMA3.4, and ESOI cover the period from 1877 to 2005; NMA3.4 has gaps in 1911–12, 1916–20, and 1946–47; and the coral proxy covers the period from 1886 to 1998. For decadal and longer-period time scales, Niño-3.4, NMA3.4, and ESOI cover the period from 1882 to 1999; NMA3.4 has gaps in 1906–25 and 1941–52; and the coral proxy covers the period from 1891 to 1992.

	NMA3.4						ESOI						Coral proxy					
	Interannual			Decadal and longer			Interannual			Decadal and longer			Interannual			Decadal and longer		
	B	A	T	B	A	T	B	A	T	B	A	T	B	A	T	B	A	T
Niño-3.4	84	97	90	93	85	89	87	91	90	86	88	85	73	75	71	73	87	79
NMA3.4							76	89	83	86	92	86	62	76	68	78	79	77
ESOI													65	72	64	64	89	70

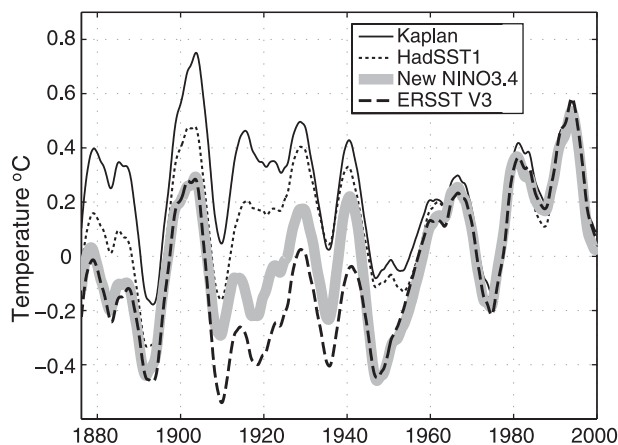


FIG. 10. Comparison between Niño-3.4 indexes from other datasets often used (Kaplan, HadISST, and ERSST V3) and the new Niño-3.4 constructed here. The series have been decadal filtered. Note that the Kaplan, HadISST, and ERSST V3 time series have been shifted vertically so that all time series coincide at the end of the record, when data are abundant. The largest differences between the indexes occur before 1955; these differences affect the general trend of the indexes.

Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4). As noted by Vecchi and Soden, the models' scenario "corresponds roughly to a doubling in equivalent CO_2 between 2000 and 2100, after which time the radiative forcings are held constant with some of the model integrations continuing for another 100–200 yr." They found that in all 22 models, as the climate warmed, the atmospheric and oceanic circulation over the equatorial Pacific resembles El Niño-like conditions in the limited sense that the Walker cell, the SLP gradient, and the thermocline tilt across the equatorial Pacific weaken, and the convection region shifts eastward.

What do observations show? Observations of the equatorial surface atmospheric pressure difference across the Pacific (Clarke and Lebedev 1996; Vecchi et al. 2006; Vecchi and Soden 2007, Figs. 9 and 11c) show that, since the 1970s when anthropogenic global warming has very likely occurred (see Fig. SPM.4 in Solomon et al. 2007), the zonal surface equatorial Pacific winds are becoming more westerly; that is, the Pacific easterlies are slowing down. In addition, during recent times, when SST data are accurate, various SST datasets agree that the SST is increasing (Fig. 10). Thus these observations are in agreement with the IPCC results and with the basic physics (Clarke 1994) that increased equatorial deep atmospheric convection results in a westerly wind anomaly (weaker easterly trade winds).

Vecchi et al. (2006) also showed, using the zonal surface equatorial difference (ESOI) that the equatorial Pacific easterlies have slowed down over the much longer period of 1861–1992. However, the SST datasets available

TABLE 3. Estimation of the trend of various indexes. The series used were monthly and, except for the coral proxy Niño-3.4, all series ran from 1877 to 2005. The coral proxy ran from 1890 to 1998. To give some idea of the chance that the overall trend just arose randomly, we must take into account that the monthly data are not independent. We do this in the same spirit as Ebisuzaki (1997) by creating 20 000 time series with exactly the same autocorrelation function as the index time series in question. In practice this is done by calculating 20 000 random-phase time series with the same power spectrum as the index. The trend of all the random time series was then calculated and compared to the value obtained for the original index. Column 3 of the table indicates the likelihood that the linear trend in column 2 could have arisen by chance. For example, the percentile entry of 94 in row 4 of column 3 indicates that 94% of the randomly calculated slopes were less than the 0.31 observed slope. In the single case where the observed slope was negative (Kaplan, first row), the entry 70 in the third column is the percentage of random slopes greater than the -0.09 observed slope.

Index	Trend [$^{\circ}\text{C}$ (100 yr) $^{-1}$]	Percentile (%)
Kaplan	-0.09	70
HadISST	0.09	80
ERSST V3	0.4	93
New Niño-3.4	0.31	94
Negative ESOI	0.30	85
NMAT3.4	0.39	97
Coral proxy	0.32	99

to them did not agree that the SST has risen in the eastern equatorial Pacific during that period (see for example the disagreement between Kaplan, HadISST, and ERSST V3 datasets over that period in Fig. 10). Our results agree with the IPCC coupled model results in that the behavior of our negative ESOI, Niño-3.4, and NMAT3.4 indexes is consistent with a slowdown of the Pacific equatorial easterlies and a rise in Niño-3.4 SST and air temperature over the period since the 1970s (Fig. 11).

Since 1877, in terms of Niño-3.4 units of $^{\circ}\text{C}/100$ yr, the trend is 0.34 for Niño-3.4, 0.45 for NMAT3.4, 0.24 for ESOI, and 0.54 for the coral proxy (Table 3). However, as already noted by other authors (e.g., Cane et al. 1997; Liu et al. 2005), the size and sign of these trends may vary depending upon the time interval that is chosen. In any case, although it is possible to estimate a linear trend over the whole record, the visual impression of Fig. 11 is not of a linear trend but rather of a negligible trend from 1880 to about the 1970s and then a rise or possibly a sudden upward jump since then. A rise since the 1970s is consistent with the rise in temperature produced by IPCC models provided anthropogenic forcing is included (see Fig. SPM.4 in Solomon et al. 2007).

7. Summary and discussion

The aim of this study was to improve our knowledge of interdecadal and longer-term variability in ENSO

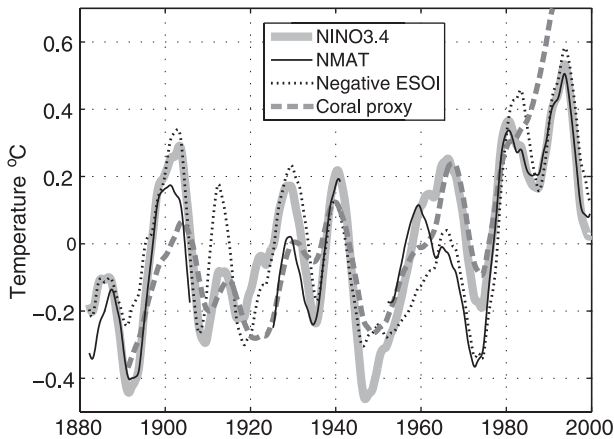


FIG. 11. Decadally filtered new Niño-3.4 (thick light gray), NMAT3.4 (thin black), negative ESOI (dotted black), and Niño-3.4 coral proxy (dashed dark gray). The NMAT, ESOI, and coral time series were converted into Niño-3.4 proxies by dividing each time series by its own standard deviation (0.23°C , 0.32 hPa , and 0.10% , respectively) and then multiplying by the Niño-3.4 standard deviation (0.23°C).

indexes. To do this we constructed three independent indexes physically related to ENSO variability: the average SST anomaly over the Niño-3.4 region, an average nighttime marine air temperature over the Niño-3.4 region, and a new ESOI. The construction of these indexes involved taking into account the uneven distribution of data in space and time and the phase locking of ENSO amplitude to the seasonal cycle. The latter offered a way to interpolate missing or doubtful data by fitting an appropriate 12-month amplitude structure to the gappy time series. These procedures were helpful in reducing noise in the monthly time series but did not affect significantly the decadal filtered time series (i.e., the decadal filtered raw time series are not very different from the weighted and fitted time series; Fig. 7).

Probably the most crucial aspect of the validation was the construction of a new ESOI. In its definition this index took into account that equatorial atmospheric turbulent surface boundary layers are deeper in the western than eastern Pacific. The index also used carefully verified atmospheric SLP data. Regarding the latter, the SLP data in the eastern equatorial Pacific close to the coast, where data density is greatest, were suspiciously noisy and data in the west also showed inconsistencies before 1950. Removing data near the eastern border and using the well-reputed Darwin SLP record instead of the average SLP in the west improved both the agreement between SLP at each end of the equator in the Pacific and also the agreement between ESOI and the new Niño-3.4.

All three indexes constructed here (Niño-3.4, NMAT3.4, and the ESOI) are much better correlated on decadal and longer time scales before 1950 than the

other Niño-3.4 indexes and SOI (Fig. 1 and Tables 1, 2). Since all three indexes are physically related but based on different data, we claim that we have improved our knowledge of decadal variability and trend of ENSO indexes in the last 130 yr. The better performance of our Niño-3.4 index compared with other Niño-3.4 indexes at interdecadal and longer time scales suggests that the interpolation schemes performed in some reconstructed SST datasets to obtain smooth-gap-free time series can be misleading when trying to assess long-term trends.

The new Niño-3.4 index is highly consistent with the ERSST V3 reconstruction (Smith et al. 2008), which shows an eastward shift of the western Pacific warm pool in recent decades (Vecchi et al. 2008). This eastward shift of the equatorial warm pool, along with the increasing westerly equatorial wind anomaly suggested by the new ESOI, is consistent with the physics (Clarke 1994) that anomalous deep equatorial atmospheric convection results in westerly equatorial wind anomalies. Much of the observed change has occurred since the early 1970s. Model results (see Fig. SPM.4 in Solomon et al. 2007) suggest that this change is due to anthropogenic forcing.

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